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On the Role of Vertical Neutral-Gas Motions in Producing Ion Convergence at E-Region Heights

WILLIAM H. HOOKE

BOULDER, COLO.
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William H. Hooke

Theoretical and observational results are used to suggest that vertical neutral-gas motions are generally important in applications of the magnetoshear theory of sporadic E, as has been recently demonstrated in a particular instance by M. A. MacLeod. It is argued that much of the vertical neutral-gas motion present at these heights must be of gravity-wave origin and shown that gravity-wave-associated vertical motions of the neutral gas should support (oppose) the action of wave-associated horizontal motions in producing ion convergence when the wave is propagating toward the east (toward the west). It is also argued that the full vector ion velocity, rather than only its vertical component, must be used in applications of the magnetoshear theory.

1. INTRODUCTION

Recent experimental work, for example that by MacLeod (1966) and Wright et al. (1967), shows that the magnetoshear theory of sporadic E, as developed by a number of researchers (Dungey, 1959; Whitehead, 1960, 1961, 1966; Axford, 1961, 1963; Axford and Cunnold, 1966; MacLeod, 1966; Chimonas and Axford, 1968), can make use of rocket-trail measurements of the neutral-gas motion at E-region heights to predict the altitudes of the relatively thin layers of enhanced electron density commonly known as sporadic-E layers or E_s layers. There are exceptions to this rule, however. In 5 of the 28 cases he studied, MacLeod (1968) found E_s layers where theory predicted a region of depleted free electron number density. Since it is assumed in the usual application of the theory that the neutral-gas motions are perfectly horizontal, he reasoned that vertical neutral-gas motions, if present, might account for the occasional discrepancies. By programming a rocket to release puffs rather than a continuous trail of TMA (trimethylaluminum), he was able to determine vertical as well as horizontal neutral-gas velocities. In the single case he reported, these vertical motions (as great as 24 m/s) were found to be the dominant contribution to ion convergence.

The present paper deals, in a general way, with the role of vertical neutral-gas motions in producing ion convergence at E-region heights. Section 2 briefly reviews the observations of such motions and the theory of their origin, both of which suggest that the motions are of turbulent or internal-gravity-wave origin. The next three sections assume, corresponding to the usual practice in applications of the magnetoshear theory, that only the vertical <u>ion</u> velocity u_{iz} is relevant, and explore the implications of including non-zero vertical neutral-gas motions in the standard equations for u_{iz}. Section 3 shows that because the ratio of ion-neutral collision frequency to ion gyrofrequency is large in the lower E region, vertical neutral-gas motions, even if small compared with the horizontal neutral-gas motions, contribute significantly to u_{iz}. Section 4 shows that vertical neutral-gas motions of internal gravity-wave origin affect u_{iz} in such a way as to favor E_s production by waves propagating in certain azimuthal directions. Such vertical motions similarly affect the wave transport of ionization (the so-called corkscrew mechanism discussed by Axford (1961, 1963) and Chimonas and Axford (1968). In particular, they change the height at which a layer is dumped by a wave (sec. 5).

Section 6 suggests, however, that in applications where the vertical neutral-gas velocities are significant, their effects on u_{iz} and hence on ion convergence may be cancelled by accompanying effects of the wave on the horizontal ion motion. As a result, the theory developed in sections 3 through 5 may be of merely academic interest. By the same token, however, section 6 suggests that sporadic-E experiments in which only the vertical wind variation is measured and wind variations in the horizontal are ignored, do not provide a definitive test of the theory. Section 7 contains concluding remarks.

2. VERTICAL MOTIONS OF THE NEUTRAL GAS AT E-REGION HEIGHTS

Few observations of neutral-gas vertical motions at E-region heights have been reported. Manning et al. (1950) found the mean winds to be horizontal to within a few degrees. They later found that while the vertical component of these mean motions was very nearly zero, it could possibly be as large as 3 m/s (Manning et al., 1954). On the other hand, they found the rms irregular vertical winds to lie between 0 and 20 m/s for the various times they attempted the relevant data reduction. They attributed these vertical motions to turbulence, but it is now believed (Hines, 1960, 1963) that a significant component of the motions considered turbulent at that time is, in fact, properly a portion of the internal-gravity-wave spectrum.

Elford and Robertson (1953) found vertical motions of relatively great horizontal extent (> 50 km) and persistence (> 1 hr) that varied between $\pm$ 6 m/s. The average angle of tilt of the wind vector from the horizontal was less than 5°. They pointed out, however, that more irregular motions of the neutral gas, even if present, would not be detected by their observational method.

Because early observations revealed the dominance of the horizontal motions of the neutral gas at E-region heights, many of the data obtained on neutral gas motions since then have been reduced routinely on the assumption that the observed motions are strictly horizontal. There are, fortunately, several exceptions:

(a) By studying the motion of filaments of rocket releases that had been blown nearly horizontal by a large wind shear, Manring et al. (1961) measured vertical neutral-gas velocities in two separate cases, finding in the first case a mean downward motion of 3.67 m/s and in the second an average upward velocity of 10 m/s.

(b) By photographic observations of chemical releases, Moseley and Justus (1967) found an rms turbulent vertical velocity at E-region heights the order of 15 m/s. The turbulence appeared to be isotropic.

(c) Murphy et al. (1967) found that the motion of a cesium cloud at 100 km, observed optically, had a 35 m/s upward component for a period of several minutes. Later motion of the cloud was deduced indirectly from ionosonde data; it apparently moved upward at 3 m/s for at least 10 min. The authors also reported optical observations of the apex of a TMA rocket trail that indicated a mean upward velocity of some 30 m/s throughout a 4-min observing period.

(d) MacLeod (1968) measured the vertical motion height profile at Arecibo in the manner described briefly in section 1. He found a rather oscillatory vertical velocity profile with an amplitude of the order of 10 to 20 m/s. He gave no information about the persistence of this profile.

(e) Wright and Fedor (1969) describe an extension of the radio drift measurement technique that permits measurements of the vertical drift of small-scale ionospheric irregularities. They tentatively equate the vertical drifts they observe with the vertical neutral-gas velocity at the irregularity height, and find E-region vertical velocities up to 15 m/s by this means.

According to theory, vertical neutral-gas motions of the order of a few centimeters per second if constant for periods of several days and if of wide spatial extent, should have profound, easily detectable atmospheric effects on temperatures (Kellogg, 1961), on airglow

(Tohmatsu and Nagata, 1963; Gadsden and Marovich, 1969), and on noctilucent cloud formation (Chapman and Kendall, 1965). The prevailing vertical motions indeed appear to be limited to this small magnitude.

Lindzen (1967) has calculated the vertical motions resulting from the thermally driven diurnal tide in an equinoctial, isothermal model atmosphere. His results show that at the equator, where the amplitude of the vertical velocity fluctuations is largest, both in absolute terms and relative to the horizontal motions, it is of the order of 1 m/s at E-region heights.

Internal atmospheric gravity waves are similar in many respects to atmospheric tides. They are characterized by smaller temporal and spatial scales, however, so that the earth's rotation and sphericity can be ignored in their description.

Hines (1960) finds that for a single plane internal gravity wave of angular frequency ω, horizontal wave number k_h, and vertical wave number k_z satisfying $4k_z^2 >> 1/H^2$, where H is the neutral-gas scale height,

$$\frac{u_z}{u_h} \doteq -\frac{k_h}{k_z} \quad . \tag{2.1}$$

The gravity-wave dispersion equation in this case reduces to Hines, 1960)

$$\frac{k_z^2}{k_h^2} = \frac{\lambda_h^2}{\lambda_z^2} \doteq \frac{\tau^2 - \tau_g^2}{\tau_g^2} \quad , \tag{2.2}$$

where $\tau \equiv 2\pi/\omega$ is the wave period and $\tau_g \equiv 2\pi C/[(\gamma-1)^{\frac{1}{2}} g]$ is the Brunt period. At E-region heights, $\tau_g \sim 5$ min and $H \sim 6$ km. Figure 1 shows the dependence of the ratio $|u_h/u_z|$ on the ratio τ/τ_g.

Hines (1960) found that waves of 12-km vertical wavelength and a period of the order of 1 to 2 hours were dominant in the meteor trail observations. Kochanski (1964) found from analysis of rocket release trails an average amplitude for the wave-associated horizontal velocities of $\sim$ 50 m/s. From the above equations one finds an average amplitude for the vertical motions associated with these dominant waves of $\sim$ 2.5 to 5 m/s. Hines (1963) noted that smaller-scale waves (with vertical wavelengths of the order of 1 km) seemed to contribute the same amount to the total wind shear as did the larger-scale, dominant waves. Thus the horizontal motions associated with the smaller-scale wave components must be $\sim$ 5 m/s. He also pointed out that such small-scale waves must have very short periods or else they would be subject to severe viscous dissipation at these heights. He concluded that their horizontal and vertical wavelengths must be comparable. The equations above tell us that in this event the smaller-scale wave components must also cause vertical motions of $\sim$ 5 m/s. Thus the smaller-scale wave components may well be the important contributors to the vertical neutral-gas motions at E-region heights.

The above numbers represent average values. Some fraction of the broad spectrum of waves present at any one time may well be creating vertical motions > 10 m/s. For example, Rosenberg (1966) states that while the mean maximum wind shear at E-region heights is 0.02 s^{-1}, the maximum wind shear at these heights is on occasion as great as 0.1 s^{-1}. A wave of 1-km horizontal and 1-km vertical wavelength, the smallest-scale wave that can survive viscous dissipation at these heights, would generate 8 m/s vertical motions in creating this shear. A wave of 10-km vertical wavelength and 100-km horizontal wavelength would generate 20 m/s vertical motions in creating this shear.

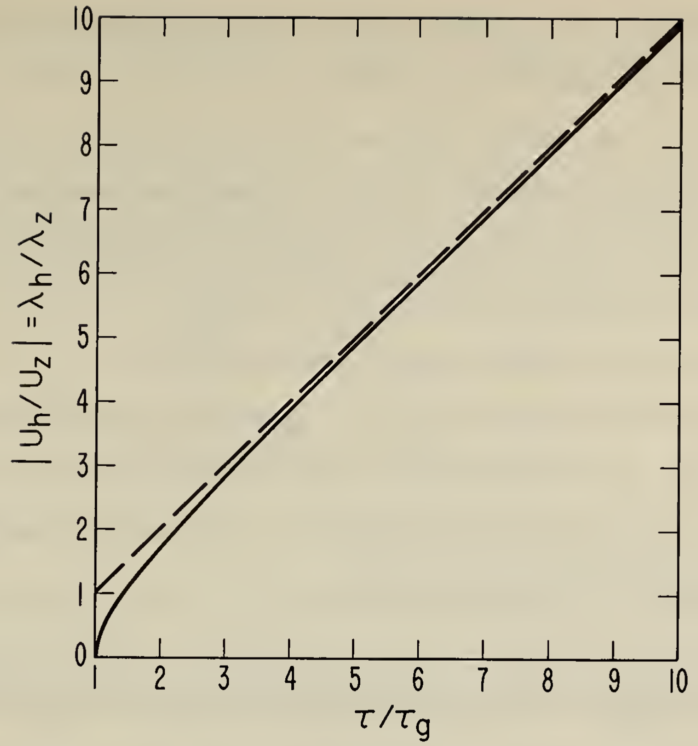

Fig. 1. The ratio of internal-gravity-wave-produced horizontal to vertical velocities as a function of the ratio of the wave period τ to the Brunt period τ_g. The solid curve is calculated from (2.1) and (2.2). The dashed curve shows the approximate result $|u_h/u_z| = \tau/\tau_g$.

Neutral-gas turbulence at low E-region heights is maintained by processes not yet fully understood, but it seems likely that the energy for its maintenance is supplied by dissipation of internal atmospheric gravity waves and the atmospheric tides (Hines, 1963, 1965). Hines (1965) estimates the rate of wave energy loss per unit mass ε to be of the order of 10^{-1} watt/kg. This estimate is a factor of 10 larger than the estimate he had used previously (Hines, 1963) to determine the order of magnitude of the turbulent velocity v_t associated with the large-scale eddies, and gives $v_t \lesssim 10$ m/s, which agrees in order of magnitude with the findings by Moseley and Justus (1967).

In conclusion, then, the largest vertical neutral-gas motions at E-region heights are probably produced by motions of small temporal and spatial scales, caused by internal gravity waves and turbulence. The gravity-wave-associated vertical neutral-gas motions should typically be ~ 5 to 10 m/s, some 10 percent of the dominant horizontal motions, though they may be larger on occasion. Turbulence-associated vertical motions of the neutral gas may be typically ~ 15 m/s. The vertical motions of the neutral gas are probably only rarely comparable to the horizontal. If the two types of motion are to be of roughly equal importance in producing ion convergence, the reason must lie in other factors, discussed in the next section.

3. VERTICAL NEUTRAL-GAS MOTIONS IN THE MAGNETOSHEAR THEORY

According to the magnetoshear theory, the mean ion velocity $\underset{\sim}{u}_i$ at any point is related to the mean neutral-gas velocity $\underset{\sim}{u}$ at that point by (MacLeod, 1966)

$$\underset{\sim}{u}_i \doteq \frac{1}{1+\rho_i^2}\left[\rho_i^2\,\underset{\sim}{u} + \rho_i\,(\underset{\sim}{u} \times \underset{\sim}{\Gamma}) + (\underset{\sim}{u}\cdot\underset{\sim}{\Gamma})\,\underset{\sim}{\Gamma}\right], \tag{3.1}$$

where $\rho_i \equiv \nu_i/\omega_i$ is the ratio of the ion-neutral momentum transfer collision frequency ν_i to the ion gyrofrequency ω_i, and $\underset{\sim}{\Gamma}$ is a unit vector directed parallel to the earth's magnetic field. Here inertial effects, electric field effects, and diffusion effects are ignored.

This equation has rarely been applied in its general form. In the usual applications it is assumed first that only the vertical ion velocity is of interest:

$$u_{iz} \doteq \frac{1}{1+\rho_i^2}\left[(\rho_i^2+\Gamma_z^2)\,u_z + \rho_i\Gamma_y u_x + \Gamma_y\Gamma_z u_y\right] , \qquad (3.2)$$

where x, y, and z are the axes of a Cartesian coordinate system with x-axis directed toward geomagnetic east, y-axis directed toward geomagnetic north, and z-axis directed vertically upward. It is then assumed that $u_z = 0$, so that

$$u_{iz} \doteq \frac{1}{1+\rho_i^2}\left[\rho_i\Gamma_y u_x + \Gamma_y\Gamma_z u_y\right] . \qquad (3.3)$$

A number of authors have pointed out that at E-region heights at low magnetic latitudes (where temperate-latitude E_s occurs most frequently), $\rho_i >> \Gamma_z$, so that if u_x and u_y are comparable, or if u_x greatly exceeds u_y,

$$u_{iz} \doteq \frac{\rho_i\Gamma_y u_x}{1+\rho_i^2} . \qquad (3.4)$$

In part this simplification results from the fact that $\rho_i > 1$ throughout the E region and that $\rho_i >> 1$ in the lower part of the E region.

Note, however, that if $\rho_i >> 1$, then u_z , when unequal to zero, may be relatively the more important contributor to u_{iz} , even if $u_z << u_x$, u_y , since the u_z coefficient in (3.2) is of the order of ρ_i^2 . Thus, for example, if $u_z \sim 0.1\, u_x$, as suggested in section 2, the contributions for u_{iz} from the two velocities should be roughly equal when $\rho_i \sim 10$, and the contribution to u_{iz} from u_z may be dominant when $\rho_i > 10$.

At the moment, ρ_i is not well known as a function of height. Figure 2 shows several estimates of this quantity. MacLeod's (1966) estimate is taken from Axford (1963, fig. 2) which is in turn based on Ratcliffe (1959, fig. 6). These estimates were apparently obtained by dividing by 50 the theoretical estimates for the electon-neutral collision frequency as a function of height calculated by Nicolet (1953). Because theoretical and observational estimates of the electron-neutral collision frequency at E-region heights still disagree (Thrane and Piggott, 1966), the validity of this profile may be somewhat questionable. Wand and Perkins (1968) calculated a ν_i height profile using the CIRA (1965) model atmosphere and formulas for the ion-neutral momentum transfer cross section obtained by Banks (1966). In this paper this ν_i height profile has been converted into a ρ_i height profile by dividing by $\omega_i \equiv 1.62 \times 10^2 \text{ sec}^{-1}$ (the value of ω_i appropriate to the E region at Boulder) and multiplying by 0.5 to convert the collisional frequency into a "frictional" frequency (see, for example, Stubbe, 1968). Wand and Perkins also measured ν_i height profiles (similarly converted into ρ_i profiles in this paper) using the incoherent scatter technique. Their results suggest significant temporal variations in this quantity (see also Wand, 1969). The Wright and Fedor (1970) ρ_i profile shown is based on Banks' (1966) results and a model atmosphere developed by Shimazaki (1967). Variations of ρ_i with ionic species, latitude, and time complicate the picture further.

Figure 2 shows that the various estimates for ρ_i differ by as much as a factor of four. Thus, according to MacLeod (1966), $\rho_i > 10$ only at heights below 100 km, while according to Wright and Fedor (1970), $\rho_i > 10$ at all heights below 104 km. The profile computed from the CIRA (1965) model atmosphere suggests that $\rho_i > 10$ at all heights below 107 km.

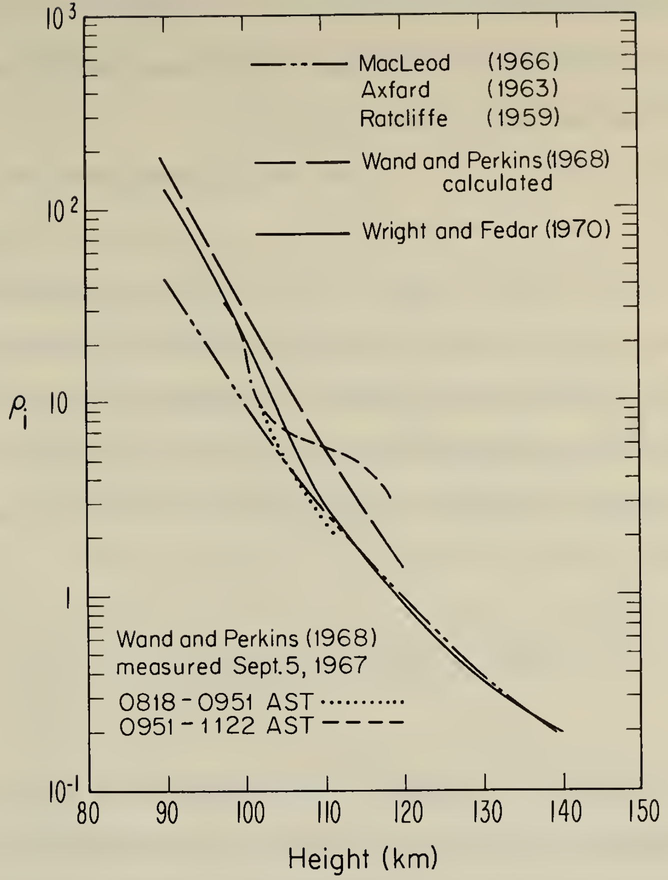

Fig. 2. Various estimates of the dimensionless parameter $\rho_i = \nu_i / \omega_i$ as a function of height.

Which ρ_i profile one uses, then, may determine the importance to be attached to vertical motions of the neutral gas in producing ion convergence. In what follows the Wright and Fedor (1970) profile is taken to be representative.

Vertical neutral-gas velocities, if typically some 10 percent of the horizontal velocities, as suggested in section 2, should then be important contributors to the vertical ion velocity at heights below about 110 km, where they may affect the latter velocity by as much as 30 percent, while they should be dominant contributors to the ion vertical velocity at heights below 104 km. In a few instances, when the vertical velocities may be an even larger fraction of the total neutral-gas velocities, they may contribute significantly to the ion velocities even at the greater heights.

Note that $\rho_i \sim 50$ in the height range of 90 to 100 km. Values of ρ_i of this magnitude imply that even small vertical neutral-gas motions of tidal origin might be as important as horizontal neutral-gas motions in producing ion convergence at these heights.

Ion convergence depends, of course, not only upon the ion velocity but also upon the variation of that velocity. The physical quantity of interest in the magnetoshear theory is the divergence of the ion flux,

$$\underset{\sim}{\nabla} \cdot (N_i \underset{\sim}{u}_i) \doteq \underset{\sim}{u}_i \cdot \underset{\sim}{\nabla} N_i + N_i \underset{\sim}{\nabla} \cdot \underset{\sim}{u}_i \ , \tag{3.5}$$

where N_i is the ion number density (equal to N_e, the free electron number density). It is known from observations that vertical gradients of ionization density in E_s layers are much greater than the corresponding horizontal gradients. Observations of neutral-gas motions also show that the dominant motions exhibit vertical variations much greater than the corresponding horizontal variations. As a result, (3.5) is usually written as (although see sec. 6)

$$\underset{\sim}{\nabla} \cdot (N_i \underset{\sim}{u}_i) \doteq u_{iz} \frac{\partial N_i}{\partial z} + N_i \frac{\partial u_{iz}}{\partial z} \ . \tag{3.6}$$

Ionization may accumulate near levels where the negative of this expression (the convergence) achieves maximum values, to an extent limited by ion diffusion and chemical loss. The ion velocity u_{iz} thus contributes to ion convergence in two well-known ways. First, it may act on a large gradient of N_i to produce convergence. Vertical and horizontal neutral-gas velocities contribute to ion convergence through this term in the same proportion as they contribute to the ion velocity itself. Thus, when vertical neutral-gas motions are an important contributor to u_{iz} itself, they are an important contributor to this term of the ion convergence.

The ion velocity may also exhibit large, rapid spatial variations, thus contributing significantly to the ion convergence through the rightmost term of (3. 6). The relative contribution of vertical and horizontal neutral-gas velocities to this term depends not only on their relative magnitudes, but also on their spatial variation. If the spatial variations of u_z and u_x are related, as when for example both motions are associated with the passage of a single atmospheric wave, then their relative contribution to the rightmost term of (3. 6) is again the same as their relative contribution to u_{iz} itself.

The situation is more complicated when all the neutral-gas motion present is of wave origin and several waves are present. Models could obviously be chosen that would lead to almost any conclusion concerning the relative importance of wave-associated vertical and horizontal neutral-gas motions in producing sporadic-E layers, and in the light of our present poor knowledge concerning details of the upper atmospheric motions, any such discussion must be speculative.

Nevertheless, one simple case is instructive. Suppose there are two internal gravity waves present, both satisfying (2. 1). Suppose, further, that the first of these waves has a 10-km vertical wavelength and a 60-min period, while the second has a 1-km vertical wavelength and a 10-min period. It follows from (2. 2) (remembering that $\tau_g \sim 5$ min) that the vertical neutral-gas motions associated with the first wave are ~ 0.1 of the corresponding horizontal motions, while the vertical neutral-gas motions associated with the second wave are comparable to the corresponding horizontal motions. Suppose in addition that the neutral-gas wind shear resulting from each of the two waves is comparable, as Hines (1963) has suggested would typically be the case (see sec. 2). Then

$$k_{z1}\, u_{h1} \sim k_{z2}\, u_{h2} \;, \tag{3.7}$$

where the subscripts 1 and 2 refer to the first and second waves respectively. Since $k_{z1} \sim 0.1\, k_{z2}$, u_{h1} must be $\sim 10\, u_{h2}$. The horizontal motions of the larger-scale wave are the dominant motions present. The wave-associated vertical motions in this case are some 10 percent of the horizontal motions, and they would probably not be noticed in the usual type of wind measurement now made.

It follows from (2.1) that

$$u_{h1} \sim 10\, u_{z1} \quad \text{and} \quad u_{h2} \sim u_{z2} \;, \tag{3.8}$$

so that while $k_{z1}\, u_{z1}$, the variation of u_{z1} in the vertical direction, is only $\sim 0.1\, k_{z1}\, u_{h1}$, just as $u_{z1} \sim 0.1\, u_{h1}$, it turns out that $k_{z2}\, u_{z2} \sim k_{z1}\, u_{h1}$, even though $u_{z2} \sim 0.1\, u_{h1}$. This suggests that in the typical case, the contribution of the small-scale vertical motions of the neutral gas to the ion convergence may be much more important than the relative magnitudes of the horizontal and vertical neutral-gas motions might suggest.

4. VERTICAL NEUTRAL-GAS MOTIONS AUGMENTING AND OPPOSING CORRESPONDING HORIZONTAL MOTIONS

Obviously, if the amplitude and phase of the vertical and horizontal motions of the neutral gas are not physically related, these motions may in general act either in concert or in opposition in producing ion convergence, to the resulting benefit or detriment of the ionospheric irregularity so produced. Even when the two components of motion are related in amplitude and phase, however, as they are when associated with the passage of a single atmospheric wave, they may still act either in concert or in opposition. Consider, for example, a single, plane internal gravity wave propagating either due east or west in an otherwise stationary atmosphere. Then (3.2) reduces to

$$u_{iz} \doteq \frac{1}{1+\rho_i^2}\left[(\rho_i^2 + \Gamma_z^2)\, u_z + \rho_i \Gamma_y u_x\right] . \tag{4.1}$$

Using (2.1), one finds that

$$u_{iz} \doteq \frac{u_x}{1+\rho_i^2}\left[(\rho_i^2 + \Gamma_z^2)\left(-\frac{k_x}{k_z}\right) + \rho_i \Gamma_y\right] . \tag{4.2}$$

Thus, whether the effects of u_x and u_z add to produce a relatively large u_{iz}, and hence a larger convergence or divergence of ionization, or partially or completely cancel to produce a small u_{iz}, and hence a small convergence or divergence of ionization, depends only upon the sign of k_x/k_z, since $\rho_i > 0$ and Γ_y (which is directed northward) is of the same sign in both hemispheres. Most internal gravity waves are thought to be energized below the region of observation and thus to have vertically upward energy propagation and vertically downward phase propagation, i.e., a negative k_z (Hines, 1960). Therefore, when

$k_x > 0$ (when the wave is propagating eastward), the effects of vertical and horizontal wave-associated motions of the neutral gas add constructively, while when $k_x < 0$ (when the wave is propagating westward), the effects cancel partly or completely. It follows that two waves, alike in all respects other than the sense of their east-west propagation, may have ionospheric effects of different magnitude (see also Hooke, 1968, 1969b).

It also follows that, when k_x / k_z is negative, u_{iz} has the same sign it would have if only the horizontal component of the neutral-gas motions were present; only the magnitude of u_{iz} relative to u_x is changed. In this case, regions considered to be regions of ion convergence based only upon measurements of the horizontal motions will then still be regions of convergence when the measurements of the vertical motions present are included. Comparisons between theory and experiment on the positions of the E_s-layer height should then agree, even though it may be the vertical neutral gas motions, traditionally ignored, that are the major contributors to the ion convergence.

When k_x / k_z is positive, however, u_{iz} may have either the same sign it would have had if only the horizontal component of the neutral-gas motions had been present, or, if the vertical neutral-gas motions are the dominant contributor to the ion convergence, the opposite sign. In the former case, the magnitude of u_{iz} will be greatly reduced; in the latter, we may expect to find discrepancies between observed E_s-layer heights and E_s-layer heights calculated on the basis of assumed purely horizontal motions of the neutral gas.

5. VERTICAL NEUTRAL-GAS MOTIONS AND THE CORKSCREW EFFECT

Axford (1961, 1963) and Chimonas and Axford (1968) have developed the subject of downward vertical ion transport by atmospheric waves. This effect, known as the corkscrew effect, may be simply described as follows:

It is well known that the wind patterns observed at E-region heights are not stationary. Instead, they progress vertically downward with time, because much of the atmospheric motion at these heights is of wave origin, and because the waves, energized from below, exhibit a vertically downward phase propagation. E_s layers tend to form in convergent regions of the wind profile at heights where the downward velocity of the individual ions of the layer matches the vertical phase velocity v_z of the waves, i.e., where $u_{iz} = v_z$, and follow the wave downward. In predicting E_s-layer heights, therefore, it is not enough to know the wind profile at a given instant; one must also know the phase velocity of this pattern. According to the usual picture (modified below) in which $u_z = 0$, the amplitude of the fluctuations in the velocity u_{iz} decreases with decreasing height, primarily because the coefficients of u_x and u_y in (3.2) decrease with decreasing height roughly as ρ_i^{-1} and ρ_i^{-2}, respectively. Below some height, the amplitude of the fluctuations in u_{iz} falls below the value of v_z, and the E_s layer can no longer follow the downward phase progression of the wave; the ionization is "dumped". (It does continue to exercise an oscillatory motion with a mean downward drift much less than the wave phase velocity, however; see Chimonas and Axford, 1968).

The contribution of vertical neutral-gas motions to u_{iz} obviously plays an important role in determining the height at which an E_s layer is dumped by the atmospheric wave responsible for its transport, especially since the relative importance of the vertical neutral-gas motions is greatest at the lowest E-region heights, where the dumping occurs. While the contributions to $\underset{\sim}{u}_{iz}$ (see (3.2)) from u_x and u_y decrease roughly as ρ_i^{-1} and ρ_i^{-2} with decreasing height, the contribution to u_{iz} from u_z remains roughly constant throughout the lower E region. Again, whether the vertical motions present tend to increase or decrease the amplitude of the u_{iz} variations - and hence to decrease or increase the dumping height - depends upon whether they act in concert with or opposition to the action of the horizontal neutral-gas motions in creating ion convergence. This in turn depends upon the direction of propagation of the waves responsible for the motion (see sec. 4).

These features are illustrated in figures 3, 4, and 5. Figure 3 shows a plot of the maximum downward velocity u_{iz} within the shear layer as a function of height computed from (3.2) under the assumption $u_z = u_y = 0$, and $u_x = -50$ m/s, and based on values for the u_x coefficient tabulated by Wright and Fedor (1970). Also shown are the dumping heights corresponding to $v_z = -2.5$ m/s and 0.25 m/s. The value $v_z = -2.5$ m/s is the vertical phase velocity of a gravity wave of 9-km vertical wavelength and 1-hour period, while the value $v_z = -0.25$ m/s is taken to correspond roughly to the vertical phase velocity of the diurnal tide; $u_x = 50$ m/s is chosen as a moderate amplitude for the wave-associated winds. Figure 3 shows that this maximum downward velocity falls below v_z, that is, the wave dumps its associated E_s layers at a 105-km height in the gravity-wave case, and at 93-km height in the tidal.

It is assumed in the construction of figure 3, in disregard of fact, that $u_z = 0$. We now examine the changes in figure 3 and the conclusions drawn from it when $u_z \neq 0$. In the construction of figure 4, the ratio u_{iz}/u_x is calculated from (4.2); the remaining assumptions remain unchanged. The value of $|k_x/k_z|$ is taken to be 0.01, that of

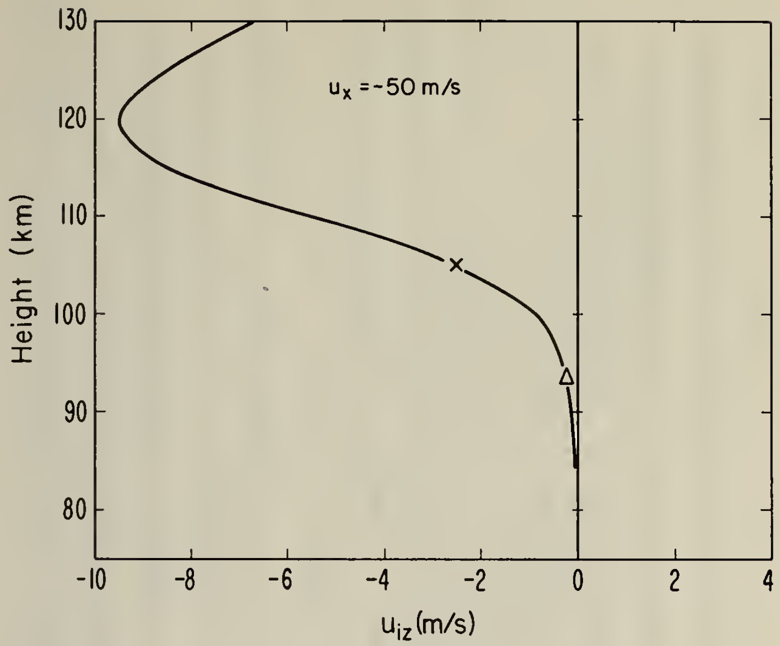

Fig. 3 The maximum downward ion velocity u_{iz} within the shear layer computed from (3.2) on the assumption $u_x = -50$ m/s, $u_y = u_z = 0$. Also shown are the dumping heights corresponding to a wave phase velocity
$v_z = -2.5$ m/s (x),
$v_z = -0.25$ m/s (Δ).

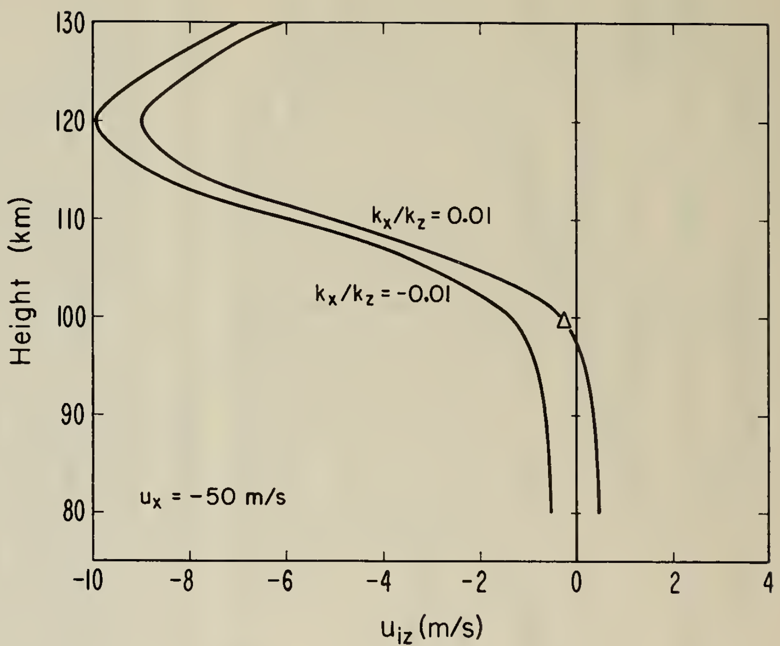

Fig. 4. The maximum downward ion velocity u_{iz} within the shear layer computed from (4.2) on the assumption $u_x = -50$ m/s, $|k_x/k_z| = 0.01$, and $v_z = -0.25$ m/s. Also shown is the dumping height (Δ) for the westward propagating wave.

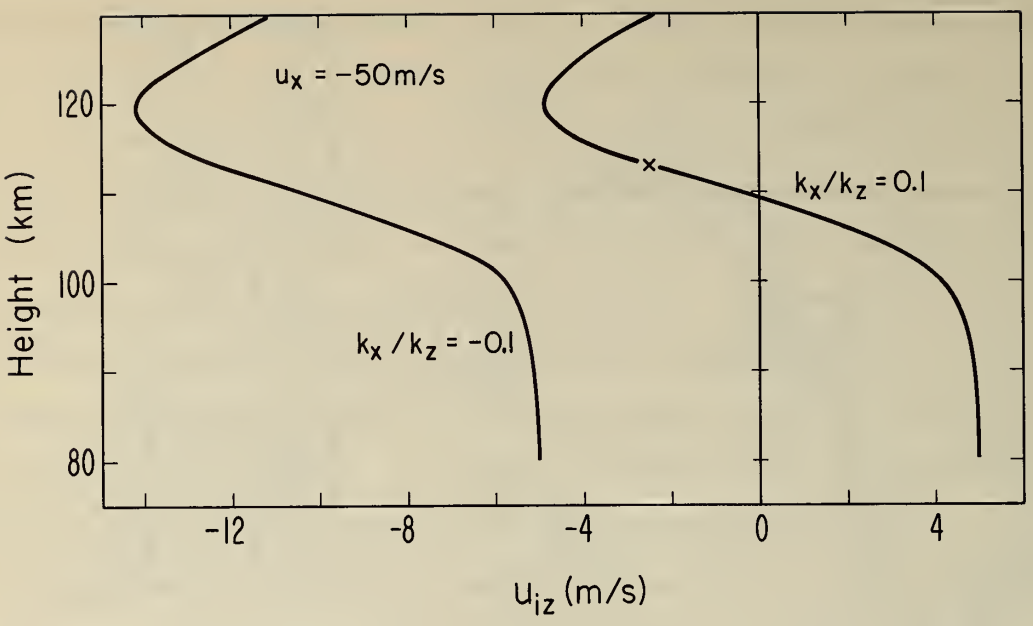

Fig. 5. The maximum downward ion velocity u_{iz} within the shear layer computed from (4.2) on the assumption $u_x = -50$ m/s, $|k_x/k_z| = 0.1$, and $v_z = -2.5$ m/s. Also shown is the dumping height (x) for the westward propagating wave.

v_z is taken to be -0.25 m/s, and that of u_x is taken to be 50 m/s. Note that for $k_x/k_z = 0.01$ (westward-propagating wave) the dumping height increases to 100 km, while there exists no dumping height for $k_x/k_z = -0.01$ (eastward-propagating wave). In fact, of course, conservation of energy requires that at heights below 100 km $|u_x|$ decreases with decreasing height (see Hines, 1960) contrary to the assumption made here, so that ultimately even an eastward-propagating wave should dump its trapped ionization. Similar comments apply to the internal-gravity-wave example depicted in figure 5. Here $|k_x/k_z|$ is taken to be 0.1, v_z is taken to be -2.5 m/s, and u_x is again taken to be -50 m/s. The westward-propagating wave dumps its trapped ionization at a height of 113 km, while the eastward-propagating wave would carry its ionization downward through all altitudes, were it not for the decrease in wave amplitude with decreasing height not taken into account here.

Thus it seems that the westward-propagating waves dump their trapped ionization because the efficiency of the magnetoshear process decreases with decreasing altitude and because wave-associated vertical motions oppose the action of the wave-associated horizontal motions in the convergence process. The wave-associated vertical motions can increase the dumping height by as much as a scale height. The eastward-propagating waves, on the other hand, dump their trapped ionization not for this reason but rather because of the decrease of wave amplitude with decreasing height.

It should be mentioned, however, that when wave-associated vertical ion velocities are important to ion transport, wave-associated <u>horizontal</u> phase velocities are also important. It is not enough that the vertical ion velocity $u_{iz} = v_z$; instead the ion motion must satisfy the condition $\underset{\sim}{k} \cdot \underset{\sim}{u}_i = \omega$ for effective transport to occur. This requires a three-dimensional treatment of the problem.

6. E_S LAYERS PRODUCED BY TRANSVERSE WAVES

The preceding sections contained the implicit assumption, usually made in applications of the magnetoshear theory, that the relevant ion velocity is the vertical component, u_{iz}, and that the other components of $\underset{\sim}{u}_i$ may be ignored. It is easy to show this assumption is a poor one. Consider the formation of an E_s layer in an originally homogeneous ionosphere. Then

$$\underset{\sim}{\nabla} \cdot (N \underset{\sim}{u}_i) \doteq N \underset{\sim}{\nabla} \cdot \underset{\sim}{u}_i \quad . \qquad (6.1)$$

If $\underset{\sim}{u}$ varies rapidly (on spatial scales $\sim$ one scale height or less), then variations in $\underset{\sim}{u}_i$ resulting from variations in ρ_i (see MacLeod, (1966) are negligible compared with variations in u_i caused by variations in $\underset{\sim}{u}$. If the neutral-gas motions are themselves of wave origin, then $\underset{\sim}{\nabla} \cdot \underset{\sim}{u} \doteq -i\underset{\sim}{k} \cdot \underset{\sim}{u}$, where $\underset{\sim}{k}$ is the wave vector, and (3.1) can be used to give

$$\underset{\sim}{\nabla} \cdot \underset{\sim}{u}_i \doteq \frac{-i}{1+\rho_i^2}\left[\rho_i^2 \underset{\sim}{k} \cdot \underset{\sim}{u} + \rho_i \underset{\sim}{\Gamma} \cdot (\underset{\sim}{k} \times \underset{\sim}{u}) + (\underset{\sim}{u} \cdot \underset{\sim}{\Gamma})(\underset{\sim}{k} \cdot \underset{\sim}{\Gamma})\right]. \tag{6.2}$$

Note that this expression contains all the components of $\underset{\sim}{u}_i$, including u_{ix} and u_{iy}, the components that are usually neglected in calculations of ion convergence.

For internal gravity waves and the atmospheric tides, however, $\underset{\sim}{k} \cdot \underset{\sim}{u} \doteq 0$ (see, e.g., Hines, 1960), so that

$$\underset{\sim}{\nabla} \cdot \underset{\sim}{u}_i \doteq \frac{-i}{1+\rho_i^2}\left[\rho_i \underset{\sim}{\Gamma} \cdot (\underset{\sim}{k} \times \underset{\sim}{u}) + (\underset{\sim}{u} \cdot \underset{\sim}{\Gamma})(\underset{\sim}{k} \cdot \underset{\sim}{\Gamma})\right]. \tag{6.3}$$

The term $\rho_i^2 u_z$ of (3.2), the term usually neglected but emphasized by MacLeod (1968) as important has been <u>cancelled</u> by other terms appearing in the other component equations of (3.1) that are also usually ignored.

This means that vertical neutral-gas motions of wave origin should in fact play only a minor role in <u>creating</u> E_s layers, although once the layer is created, they may play an important role in <u>intensifying</u>, <u>maintaining</u>, or <u>destroying</u> it. Even in this case, however, considering u_{iz} and neglecting u_{ix} and u_{iy} implies keeping terms of the order of $[\rho_i^2/(1+\rho_i^2)]u_z\, \partial N_i/\partial z$ and neglecting terms of the order of $[\rho_i^2/(1+\rho_i^2)]\, u_x\, \partial N_i/\partial x$. It is not obvious, a priori, that this procedure is at all legitimate. In fact, it would appear that the opposite is so, and that any assumption such as the above should be justified empirically on a case by case basis. J. D. Whitehead (private communication) has raised similar questions.

As a result, the development of sections 3 through 5 may be of only academic interest. Unfortunately, however, this cancellation also implies that to specify the neutral-gas wind field at E-region heights for comparison of magnetoshear theory with observed E_s-layer heights and intensities, puffs giving the height variations of all three wind components are insufficient. The horizontal variations of all three wind components are also required.

7. CONCLUDING REMARKS

The work presented above prompts the following conclusions:

(a) The limited observations and theory available suggest strongly that vertical neutral-gas motions at E-region heights are of amplitude sufficient to be important in the study of sporadic E.

(b) These vertical motions may act either in concert with, or in opposition to, the action of horizontal motions of the neutral gas in producing ion convergence, not only when the state of E-region motion is chaotic but even in cases when the components of motion are associated with a single-wave system. They may therefore affect not only the magnitude but also the sense, or sign, of the ion convergence, changing the E_s-layer intensity, E_s-layer height, and the dumping height. It appears that eastward-propagating waves should be more successful than westward-propagating waves in producing E_s layers.

(c) There is theoretical reason to expect that the full magneto-shear equation (3.1) and not just the vertical component (3.2) of that equation, together with full knowledge of the state of E-region neutral-gas motion, are required to obtain a consistent agreement between the magnetoshear theory and observation. Such discrepancies as now exist appear to result from an incomplete specification of neutral-gas motion rather than from any basic shortcomings of the magnetoshear theory itself.

Since eastward-propagating waves appear to be more successful than westward-propagating waves in creating E_s layers, one might conclude that most E_s layers should be observed traveling eastward. There is some evidence that the opposite is the case (Egan and Peterson, 1962). This may pose a serious problem for E_s theory.

The observations, however, reflect the outcome of an interplay among several competing factors. For instance, Hines (1963) and Hines and Reddy (1967) show that wind-shear filtering may cause considerable anisotropy in the directions of wave propagation at ionospheric heights. Perhaps in some cases, only westward propagating waves, or only eastward-propagating waves, are permitted at these altitudes. Then any E_s "patches", or "clouds" observed must be propagating in that direction. The rates of photochemical processes in E_s layers are affected by atmospheric waves in a way dependent upon the direction of wave propagation (Hooke, 1969a; 1969b); this may also influence the preferred direction of E_s "patch" or "cloud" travel.

Moreover, T. M. Georges (private communication) has noted that the HF ground backscatter observations of Egan and Peterson suggest that E_s patches move circumferentially, indicating an observational bias of some kind.

Finally, the dominant atmospheric tides (which are also likely to be the principal contributors to E_s) propagate only westward.

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